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OLAVI VUORELAINEN

Thermal conditions in the ground from the viewpoint of foundation work, heating and plumbing installations and draining

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THERMAL CONDITIONS IN THE GROUND FROM THE VIEWPOINT OF FOUNDATION WORK, HEATING AND PLUMBING INSTALLATIONS AND DRAINING

OLAVI VUORELAINEN

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PREFACE

Knowledge of thermal conditions in the ground is necessary for the planning and performing of various kinds of building work. In house-building activities, the depth of the permanently unfrozen ground (the course of the $0\,^{\circ}C$ isotherm) has to be known; this applies also to road, aerodrome and other construction work. Various kinds of pipelines, such as those for water mains, sewerage and drainage, have to be laid out at a depth where the ground will not freeze. Calculation of the heat losses in remote heating ducts requires knowledge of the thermal conditions in the ground. Heating systems based on the heat pump principle mostly employ the heat stored in the ground for their heat source.

The present investigation was instigated by a study of the floor heating of buildings erected upon a concrete slab poured directly on the ground, which the author has carried out in the State Institute for Technical Research in Finland.

1. FACTORS AFFECTING THE TEMPERATURE OF THE UPPER GROUND LAYERS

The field of temperature distribution in the ground is determined by the physical quality of the ground and by the entering and escaping quantities of heat. The ground obtains its entire heat from the sun by direct solar radiation (S_s), diffuse, scattered radiation from the atmosphere (S_h) or counterradiation from the atmosphere (S_v). Owing to the geothermal gradient, about 0,03 °C/m in magnitude, a quantity of heat amounting only to about 0,06 kcal/m²h flows from the interior of the earth to its surface [1]. This heat quantity is negligible with respect to its effect upon the temperature of the upper ground layers.

On the other hand the ground immediately loses part of the received radiant energy in the form of radiation reflected by the soil surface (S_r). According to Ångström's [2] and Lunelund's [3] investigations, the proportion of reflected radiation has the following magnitude, depending on the quality and covering (vegetation; snow cover) of the ground:

Quality of soil surface	Reflected quantity of heat, per cent. (Albedo)
Grey sandy plain	12 ··· 26
Black earth	12 ··· 14
Granite rock	12 ··· 18
Cereal fields, on an average	16
Grass plain	25 ··· 26
Forest	10 ··· 18
Pure snow cover	81 ··· 85
Old snow cover	42 ··· 70

The reflected quantities of heat vary greatly according to the quality of the ground.

Part of the radiant energy retained by the ground is continuously given off as radiation into space; the soil surface acts approximately in the manner of a black body giving off long-wave thermal radiation (S_u).

The difference of the heat quantities received and given off by the ground, the radiation balance (S_t), determines the temperatures of the upper ground layers (0 ··· 10 m) and the magnitude of the heat quantities stored in the ground.

$$S_t = (S_s + S_h + S_v) - (S_r + S_u)$$
$$= (S_s + S_h - S_r) - (S_u - S_v) \qquad (1)$$

The annual quantities of radiant energy in Finland are, according to Keränen [4]:

S_s = 446 Mcal/m² $\qquad$ $S_s + S_h - S_r$ = 640 Mcal/m²

S_h = 331 " $\qquad$ $S_u - S_v$ = 429 "

S_r = 137 " $\qquad$ S_t = 211 "

During the warm season, the ground receives more heat in the form of radiation than it gives off during the whole 24-hour day, i.e., the heat balance (S_t) of the ground is positive and the temperature of the surface layers increases.

According to Keränen [4], the monthly thermal balance values in South-Finland are:

S_t	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII	I ··· XII
Mcal/m²	-29	-23	-2	26	61	95	82	47	16	-12	-23	-27	211

The thermal balance is positive during the time from April to September.

The quantity of radiant energy that remains available on the ground surface (S_t) is consumed to heat the upper ground layers (Q_m), to heat the adjacent air (Q_i) and to evaporate water and to melt ice Q_h [5], [6], [7]:

$$S_t = Q_m + Q_i + Q_h \qquad (2)$$

Heating of the ground itself requires only a very small fraction of the heat quantities available on the soil surface. According to Franssila's investigations [8], only about 5 % of the radiation balance on a warm summer day remain in the ground.

A remarkable part of this radiant energy is transferred to the air by the effects of turbulence and by conduction. Great quantities of heat are consumed to evaporate water, particularly in windy weather [9]. On the other hand the ground receives heat from warm air masses and from water condensing on its surface, re-

leasing its latent heat. Rain will either heat or cool the ground, depending on the temperature of the rain water. In Finland, rain usually increases the ground temperature.

According to the investigations carried out by Singer [10] in Munich, the temperatures in the upper ground layers deviate upward from the mean the more, the greater the amount of rainfall during the preceding summer. This is mainly due to increase in thermal conductivity of the soil with increasing moisture. According to measurements by Homen [11] at Mustiala, the temperature of the surface layers of the ground is 0,5 ⋯ 3,0 °C higher in rainy summers than in dry summers.

The annual range of variation of the ground temperatures and the magnitude of the heat quantity stored in the ground depend further on the soil quality, its moisture conditions, density, specific heat, etc. The heat transfer from the surface layers to greater depths is controlled by the thermal conductivity of the soil and by the temperature gradient in the ground. The heat stored in the ground is at its maximum in the beginning of September, when the heating period of buildings begins, and it is lowest in the beginning of April [12]. In the intervening period the ground gives off the heat stored during the summer. Owing to its high thermal insulating capacity, snow cover delays the cooling of the ground. During winters abounding in snow the ground remains unfrozen up to its surface, owing to the heat stored in the ground.

2. THERMAL CONDUCTIVITY AND TEMPERATURE CONDUCTIVITY OF THE SOIL, AND THEIR DETERMINATION

For the mathematical calculation of the temperature distribution in the ground it is necessary to know a physical constant of the soil substance, its temperature conductivity a:

$$a = \frac{\lambda}{\varrho \cdot c} \tag{3}$$

This quantity is dependent on three factors: the thermal conductivity λ, the volume weight ϱ, and the specific heat c.

According to Kersten's investigations, the thermal conductivity of the soil is dependent on its moisture φ (in per cent. of the dry weight) and its dry volume weight ϱ_0 by the formula,

- for coarse soil types:

$$\lambda = (0,087 \log \varphi + 0,05) \cdot 10^{0,624 \cdot \varrho_0}$$

- for fine soil types: (4)

$$\lambda = (0,115 \log \varphi - 0,025) \cdot 10^{0,624 \cdot \varrho_0}$$

The constants in the formulas have been determined by the said author on the basis of a material comprising several thousand λ-determinations under varying conditions of moisture, volume weight and temperature. The formulas were calculated for a mean temperature of +4,5°C and the maximum deviation of the observed values from those calculated by the formulas were within 25 %, when 1 % < φ < 25 % and 1400 kg/m^3 < ϱ_0 < 2200 kg/m^3 [13].

As a rule, the dry volume weights of the different soil types vary within narrow limits, according to the investigations of Watzinger, Kersten and Saare [13].

Most soil types in the dry state have, according to Pfaundler [14], a specific heat, which is very accurately equal to 0,2 kcal/kg °C, or one fifth of the specific heat of water, if the soil is free of humus substances. In highly humous soils, the specific heat may be as high as 0,4 kcal/kg °C.

The heat content of moist soil can be computed from the specific heats of the dry soil and of water by the following formula:

$$C_w = (\varrho - w) \cdot C_k + w \cdot C_v \tag{5}$$

C_w Heat content of the moist soil [kcal/m³ °C]

ϱ Volume weight of the moist soil [kg/m³]

w Water content of the soil [kg/m³]

C_k Specific heat of the dry soil [kcal/kg °C]

C_v Specific heat of water [kcal/kg °C]

In Table 1, values of the dry volume weight, moisture content, thermal conductivity, heat capacity and temperature conductivity of the soil types occurring most commonly on building sites have been compiled, arranged by the temperature conductivity of the soil. The values are results from measurements carried out by Kersten in Alaska, E. Saare and C. G. Wenner in Sweden and Watzinger in Norway. In their measurements, Saare and Wenner have employed the non-stationary probe method, the soil being in undisturbed natural state. Watzinger has measured values of the thermal conductivity by a stationary method, the soil under investigation being placed between a heated and a refrigerated plate and subjected to load. Kersten's method is otherwise the same but in his apparatus a cylinder replaces the heated plate and another, annular cylinder the cooled plate [13]. The values obtained by the non-stationary method in the natural state are as a rule about 10 % in excess of those obtained by the stationary method. A family of graphs plotted on the basis of these values, showing the dependence of the temperature conductivity of the soil types in question on their moisture, is presented in Fig. 1.

In the nature, temperature conductivity is not a constant even if one and the same soil type is considered but it varies according to the moisture content and, in some degree, also to the temperature of the soil. At low moisture contents, a relatively small change in water content as much as doubles the temperature conductivity. This has also been shown by Keen and Patton [9] in their investigations.

The moisture in the ground varies considerably with the seasons and according to the abundance of rainfall, with consequent variations in temperature conductivity is highest in the spring and autumn, when the ground is wet. In the summer, when the ground is usually dry, it has a low temperature conductivity.

Table 1

Dry volume weight, moisture content and thermal constants of various soil types

Soil type	Dry volume weight, kg/m^3	Moisture, % of dry weight	λ Mcal/m h °C	$c\varrho$ Mcal/m^3 °C	a m^2/h	Author
Group I_1, a = 0,0022 m^2/h						
Natural esker gravel	1550	14,8	1,12	522	0,00215	Watzinger, Norway
Gravel	1760	1,9	0,87	385	0,00226	Kersten, Alaska
Sandy gravel	1800	14,9	1,36	628	0,00216	Watzinger, Norway
Gravelly medium coarse sand	1920	1,3	0,79	409	0,00195	Kersten, Alaska
Medium coarse sand	1580	4,0	0,84	379	0,00222	Saare, Sweden
Fine sand	1760	13,9	1,19	596	0,00200	Kersten, Alaska
Medium fine sand	1610	9,6	1,12	477	0,00235	Saare, Sweden
Very fine sand	1670	22,8	1,67	714	0,00234	Saare, Sweden
Sandy and fine sandy till	1930	3,6	1,11	455	0,00244	Kersten, Minnesota
Fine sandy till	1780	13,6	1,43	598	0,00239	Saare, Sweden
Very fine sandy silt	1750	15,5	1,46	621	0,00235	Kersten, Alaska
Clayey and fine sandy silt	1850	13,9	1,46	627	0,00233	Watzinger, Norway
Very heavy clay	1500	17,0	1,03	550	0,00187	Kersten, Alaska
Group II_2, a = 0,0032 m^2/h						
Gravel	1910	3,2	1,45	446	0,00325	Kersten, Alaska
Gravelly medium coarse sand	1720	10,7	1,79	528	0,00339	Kersten, Alaska
Medium coarse sand	1690	11,0	1,69	524	0,00323	Kersten, Alaska
Sandy and fine sandy till	1910	13,6	2,03	642	0,00316	Kersten, Minnesota
Fine sandy till	2090	7,5	1,82	575	0,00317	Saare, Sweden
Clayey till	1810	16,8	1,99	666	0,00299	Saare, Sweden
Clayey and fine sandy till	1750	15,4	2,07	620	0,00334	Saare, Sweden
Group III_3, a = 0,0047 m^2/h						
Gravel	2030	3,0	2,05	467	0,00439	Kersten, Alaska
Sandy and fine sandy till	2190	6,3	2,34	576	0,00406	Kersten, Minnesota

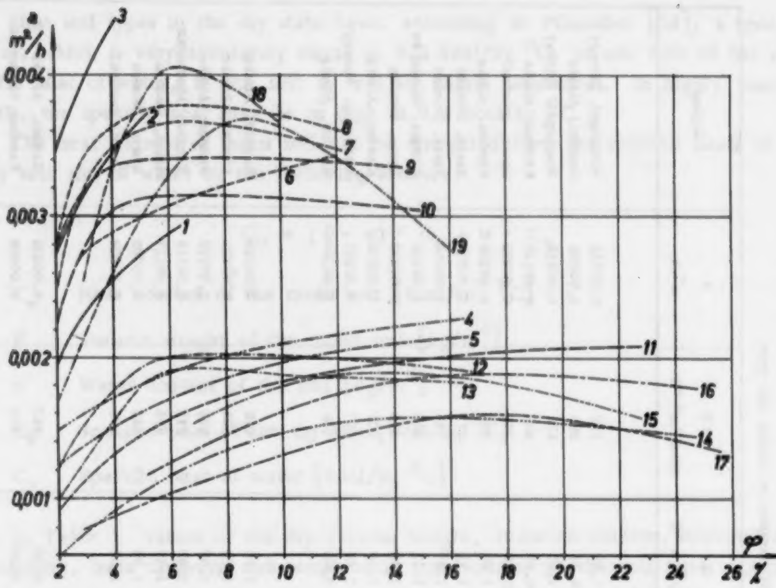

Fig. 1 Thermal conductivity of different soil types in dependence of moisture content (% of dry weight).

Key to the curves:

Gravel (dry volume weight 1760 ± 90 kg/m³),	Kersten, Alaska
" (1920 ± 40 kg/m³),	" "
" (2055 ± 75 kg/m³),	" "
Sandy gravel (1800 kg/m³),	Watzinger, Norway
Natural esker gravel (1550 kg/m³),	" "
Gravelly medium coarse sand (1716 ± 16 kg/m³),	Kersten, Alaska
" " " " (1800 ± 20 kg/m³),	" "
" " " " (1926 ± 5 kg/m³),	" "
Fine sand (1792 ± 8 kg/m³),	" "
" " (1680 ± 30 kg/m³),	" "
" " (1580 kg/m³),	Watzinger, Norway
Very fine sand (1763 ± 7 kg/m³),	Kersten, Alaska
" " " (1640 ± 10 kg/m³),	" "
Fine sandy silt (1462 kg/m³),	" "
" " " (1633 kg/m³),	" "
Very heavy clay (1498 ± 8 kg/m³),	" "
" " " (1337 ± 50 kg/m³),	" "

Sandy and fine sandy till (2116 $\pm$ 36 kg/m^3), Kersten, Minnesota
" " " " " (1920 $\pm$ 10 kg/m^3), " "

In addition to moisture, also the annual variation of the temperature in the soil affects its temperature conductivity. If one excludes the surface layer of 0,5 m depth, the soil temperature in South-Finland may vary only within the limits of -5 ··· +16 °C (Figs. 5, 6 and 7). At the upper limit of this temperature range the thermal conductivity of the soil is at its most 5 % in excess of that at the lower limit. The temperature conductivity increases with increasing depth because the density of the soil increases. At greater depth, where the moisture is less variable than in the upper layers and the ground does not freeze, more reliable values are obtained on determining the temperature conductivity. As a result of the numerous factors affecting the thermal conductivity and temperature conductivity of the soil, calculations performed with a constant value for the temperature conductivity are only approximate.

When the ground freezes, its thermal conductivity increases, as the thermal conductivity of ice, λ = 1,94 kcal/m h °C, is considerably higher than that of water, λ = 0,504 kcal/m h °C. However, coarse soil becomes more porous when it freezes, and this in turn reduces the thermal conductivity. For this reason the thermal conductivity of moist soil (below capillary saturation) changes but slightly when the ground freezes. In the state of capillary saturation the thermal conductivity of the soil (not clay) increases by 10 ··· 15 % and sometimes even by 20 ··· 30 %, when the ground freezes completely [15].

The heat content of ice, 0,45 kcal/dm^3, is less than half the quantity of heat held by the same volume of water. The heat capacity will therefore decrease when the soil freezes; since at the same time the thermal conductivity increases, the temperature conductivity will also increase. The increase in temperature conductivity caused by freezing does not attain remarkable magnitude until in very wet soil containing more than 20 % water [16].

3. THE TEMPERATURE FIELD IN FREE SOIL

a) Calculation of the temperature field

The temperature of free soil varies periodically in accordance with the seasons; the daily temperature variation is superimposed upon the annual variation. In this connection we are only concerned with the annual temperature wave.

Within a small area the soil surface may be considered to be a plane. The isotherms in the ground are then planes as well, and the heat flow is perpendicular to them. As there is no heat flow in the lateral direction, the function representing the temperature distribution in the ground, $T(z,t)$, will satisfy a one-dimensional differential equation

$$\frac{dT}{dt} = a\,\frac{d^2T}{dz^2} \tag{6}$$

The annual temperature variation at soil surface, $T(0,t)$ (boundary condition), can be represented by means of a Fourier series consisting of sine and cosine terms.

$$T(0,t) = \sum_{0}^{\infty} A_n \cos n\omega t + B_n \sin n\omega t\,, \tag{7}$$

where

$$\omega = \frac{2\pi}{t_o} \quad \text{and}$$

$$t_o = 365 \quad \text{days}$$

The temperature at different depths in the ground is found by solving the differential equation (6) under the boundary condition (7):

$$T(z,t) = \sum_{0}^{\infty} A_n e^{-\sqrt{\frac{n\omega}{2a}}z} \cos n\omega\left(t - \frac{z}{\sqrt{2an\omega}}\right) + B_n e^{-\sqrt{\frac{n\omega}{2a}}z} \sin n\omega\left(t - \frac{z}{\sqrt{2an\omega}}\right) \quad (8)$$

The coefficients A_n and B_n are determined by the boundary condition $T(0,t) = T(0,t)_{max}$ when $t = 0$.

For the coefficient A_o in equation (8) let us write $A_o = T_o$ (annual mean temperature). Equation (8) becomes then

$$T(z,t) = T_o + \sum_{1}^{\infty} e^{-\sqrt{\frac{n\omega}{2a}}z} \left[A_n \cos n\omega\left(t - \frac{z}{\sqrt{2a\,n\omega}}\right) + B_n \sin n\omega\left(t - \frac{z}{\sqrt{2a\,n\omega}}\right) \right] \quad (9)$$

The annual variation is thus obtained as the sum of a number of temperature waves. The amplitude of each wave becomes less with increasing depth and it is subject to phase lag at the same time. Each wave has its own damping factor, wavelength and velocity of propagation in the ground.

$$\text{Wavelength:} \quad L_n = 2\pi\sqrt{\frac{2a}{n\omega}} \quad (10)$$

and

$$\text{Velocity of propagation:} \quad v = \sqrt{2\,a\,n\omega} \quad (11)$$

The daily wave penetrates into the ground at higher velocity than the annual wave, which has a lower frequency. On the other hand, the daily temperature wave dies away very rapidly, extending to about 60 cm depth; it can thus be neglected when temperatures in ground layers at greater depth are being determined.

In order to calculate the annual variation of the temperatures in the ground it is necessary to know the temperature conductivity a of the soil type in question and the temperature function at soil surface, $T(0,t)$.

b) The temperature function at soil surface

No adequate data are available concerning the annual variation of the soil surface temperature but the air temperature at 2 m height from the soil surface is accurately known on the basis of meteorological observations in about 150 localities in Finland [17]. The daily means of the air temperature have therefore been used in the subsequent calculations instead of the daily mean soil surface temperatures.

Little investigation has been carried out in elucidation of the question how closely associated the variation of air temperature is with the soil surface temperatures. It is also a fact that the soil temperature at its very surface is exceedingly difficult to measure.

W. Kreutz and M. Rohweder [18] have studied this question by measuring the soil temperatures at 5, 10, 15 and 20 cm depth and calculating on their basis the soil surface temperatures. According to their experiments, the temperature of sandy soil follows the air temperatures measured at 5 cm height within about 1 °C. In Finland, measurements of temperatures at varying depth in the soil have been carried out by the observatory at Jokioinen in 1958 ··· 1959. Fig. 2 shows the annual variation of the soil temperature by daily means at 1 cm depth and the corresponding mean air temperatures, as measured in a meteorological hut at 2 m height above soil surface. As a rule, the temperature measured at 1 cm depth in the soil can be considered as representing the soil surface temperature within a few tenths of 1 °C. The temperature graphs reveal that the daily mean soil surface and air temperatures parallel each other very well. However, the soil temperature is usually on an average 1 °C higher than the air temperature during the snow-free period. In the spring and early summer, when temperature is ascending, the daily mean of the soil surface temperature may rise 3 to 4 °C above the mean air temperature, while in the autumn, when the temperature is on the downgrade, the soil surface temperature is at times 2 to 3 °C lower than the daily mean of the air temperature. In the winter, when the snow forms a heat-insulating layer upon the ground, the soil temperature remains close to the freezing point, within 1 °C. If the soil surface is kept free of snow, the daily mean of the soil surface temperature follows that of the air temperature also in wintertime.

Fig. 3 shows the annual variation of the daily mean air temperatures in Helsinki, as means of the ten-year period 1945 ··· 1954 (Curve No. 1). Immediate Fourier analysis of this curve is rendered inaccurate by the sudden jumps occurring in the curve. It was therefore smoothed by computing ten-day means and constructing from them the final temperature graph (Curve No. 2), using the three-point adjusting method. The mathematical equation of the curve was determined by Fourier analysis, determining the origin so that at the time 17. VII, 0,00 hours $t = 0$ and $T(0, t) = T_{max} + 5,48\,^{\circ}C$, because the annual mean temperature of the ten-year period in question was 5,48 °C. In the coordinate system defined in this manner,

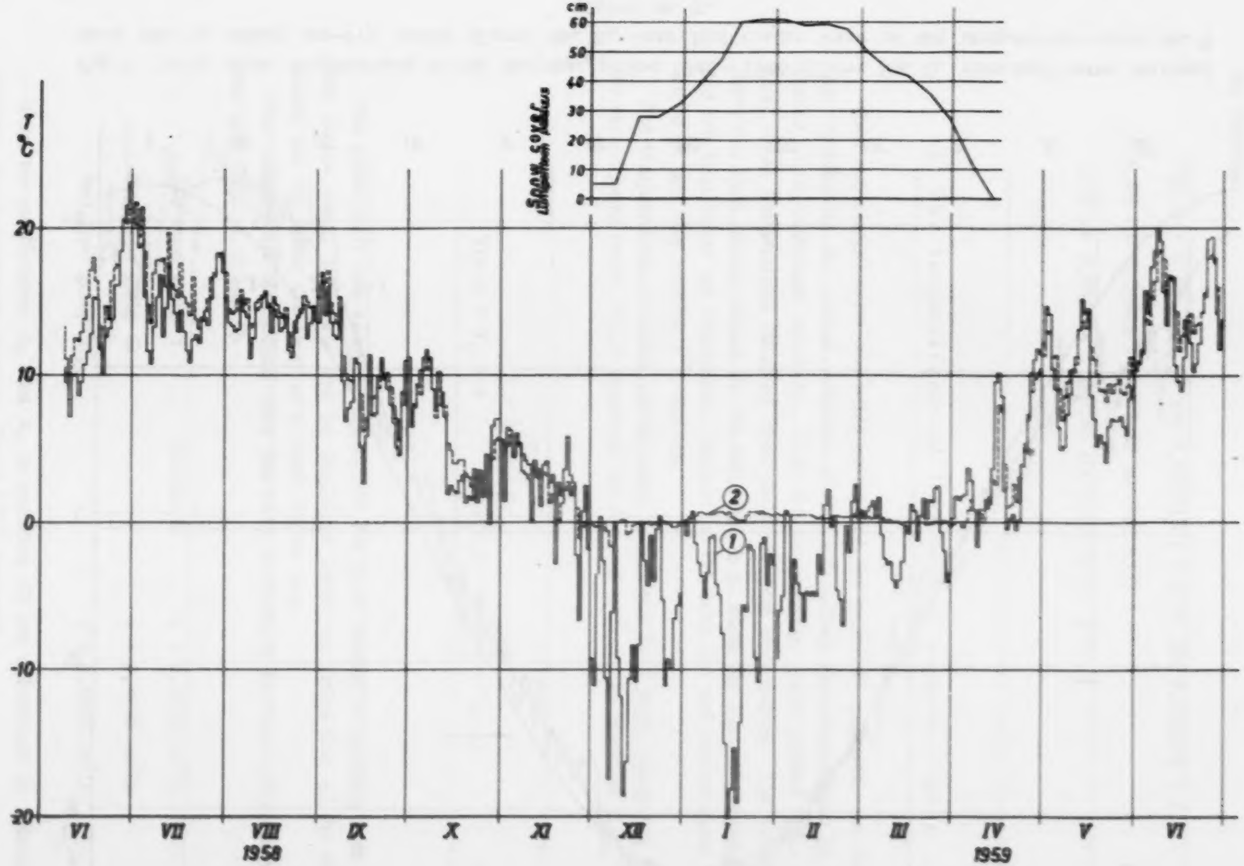

Fig. 2 Annual variation of air temperature (Curve No. 1) in a meteorological hut 2 m above soil surface, of soil surface temperature (Curve No. 2) at 1 cm depth, and thickness of snow cover in 1958 – 1959, according to measurements at the observatory of Jokioinen.

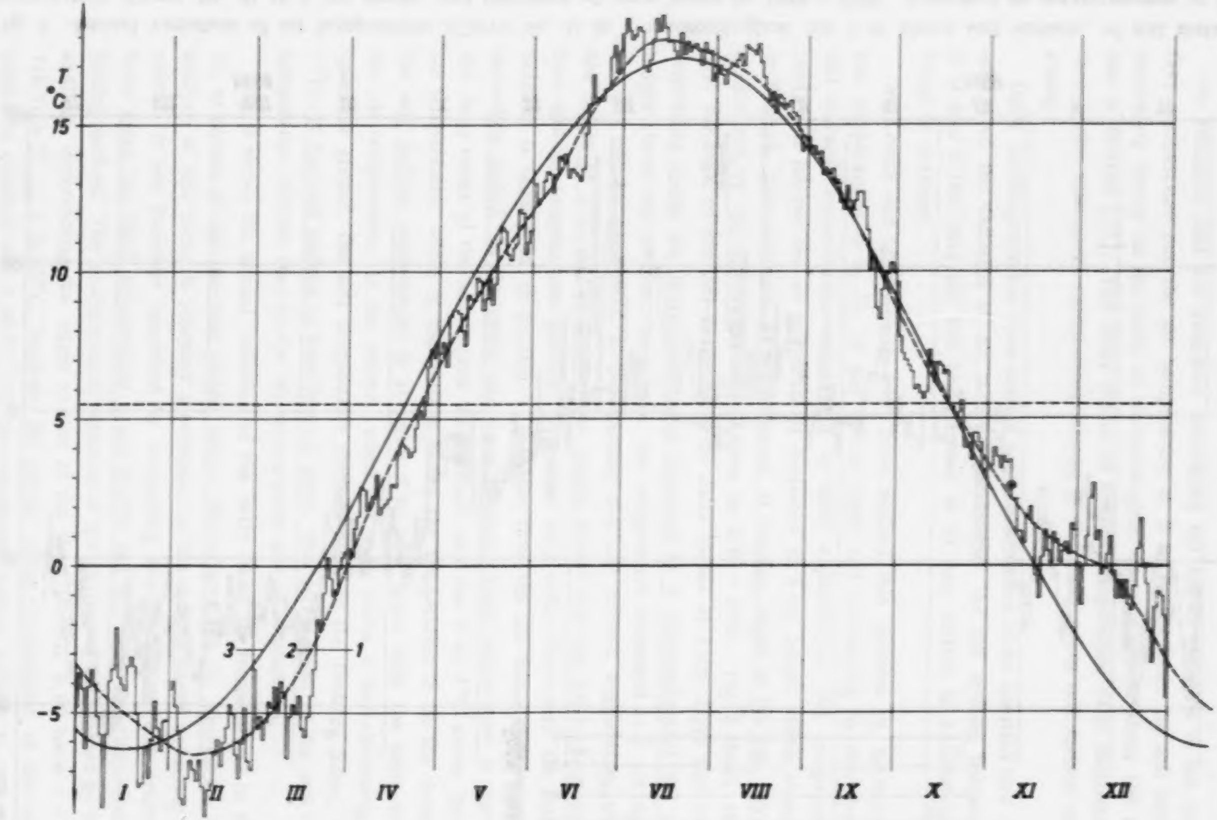

Fig. 3 Daily mean temperatures of the ten-year period 1945 – 1954 (Curve No. 1), smoothed curve obtained from this by taking ten-day means (Curve No. 2), and first cosine wave of the temperature curve No. 2 (Curve No. 3).

the variation of the mean air temperature in terms of time is thus represented by the function

$$T(0,t) = 5,48 + 11,68 \cos 0,000717\ t + 0,90 \cos 0,001434\ t + 0,06 \cos 0,00215\ t \cdots + 1,29 \sin 0,000717\ t - 1,04 \sin 0,001434\ t + 0,85 \sin 0,00215\ t \cdots\ [^{o}C.] \tag{12}$$

c) The temperatures of free soil at various depths

When soil temperatures are being calculated for technical purposes, the function representing the annual variation of the soil surface temperature can be replaced, with sufficient accuracy, with the temperature graph plotted from the daily mean temperatures obtained from meteorological observations. The calculations will then always remain on the safe side, e.g. when the limit of ground frost and the heat losses are calculated, because at the most the soil temperatures obtained in this way will be 1^{o}C too low.

The temperatures at various depths under bare ground have been calculated from the temperature curve T(0,t) measured for the soil surface, by equation (13)

$$T(z,t) = T_o + e^{-\sqrt{\frac{\omega}{2a}}\, z} \left[T\left(0, t - \frac{z}{\sqrt{2a\omega}}\right) - T_o \right] \tag{13}$$

The phase shift and damping coefficient has been chosen to conform with the first cosine wave; this wave has an amplitude about ten times that of the other waves and the latter die away at a much higher rate.

The temperature distribution field has been determined for three different values of a (Figs. 4, 5 and 6).

Soil type group I	(Table 1)	$a = 0,0022\ m^2/h$
Soil type group II		$a = 0,0032\ m^2/h$
Soil type group III		$a = 0,0047\ m^2/h$

When the coefficients A_n and B_n in equation (9) are determined by harmonic analysis, the temperature function T(z,t) [^{o}C] becomes

$$T(z,t) = 5{,}48 + 11{,}68\ e^{-\sqrt{\frac{\omega}{2a}}z} \cos 0{,}000717\left(t - \frac{1}{\sqrt{2a\omega}}z\right)$$

$$+ 0{,}90\ e^{-\sqrt{\frac{\omega}{2a}}z\sqrt{2}}\ .\ \cos 0{,}001434\left(t - \frac{1}{\sqrt{2a\omega}}\ \frac{z}{\sqrt{2}}\right)$$

$$+ 1{,}29\ e^{-\sqrt{\frac{\omega}{2a}}z}\ ,\ \sin 0{,}000717\left(t - \frac{1}{\sqrt{2a\omega}}z\right) \qquad (14)$$

$$- 1{,}04\ e^{-\sqrt{\frac{\omega}{2a}}z\sqrt{2}}\ ,\ \sin 0{,}001434\left(t - \frac{1}{\sqrt{2a\omega}}\ \frac{z}{\sqrt{2}}\right)$$

$$+ 0{,}85\ e^{-\sqrt{\frac{\omega}{2a}}z\sqrt{3}}\ ,\ \sin 0{,}00215\left(t - \frac{1}{\sqrt{2a\omega}}\ \frac{z}{\sqrt{3}}\right)$$

where the time t has been expressed in hours and $t = 0$ corresponds to July 17th, 00,00 hours. The constants occurring in this formula have been tabulated in Table 2.

Table 2

a m^2/h	$\sqrt{\frac{\omega}{2a}}$ 1/m	$\frac{1}{\sqrt{2a\omega}}$ h/m
0,0022	0,404	563
0,0032	0,335	467
0,0047	0,275	384

On the soil surface ($z = 0$) the deviation of the temperature values computed by this formula from the annual, adjusted mean value curve is no more than $1\,^{o}C$ at any point (Figs. 4, 5 and 6).

The amplitude of the annual temperature curve becomes less with increasing depth, and progressive phase shift of the wave occurs at the same time. The velocity of propagation v, wavelength L, and the depth z at which the temperature variation range is $1\,^{o}C$ are seen in Table 3 for the above-mentioned soil types. For instance, in soil type group 1 (Fig. 4) the amplitude, at soil surface $12{,}2\,^{o}C$, is $8{,}2\,^{o}C$ at 1 m depth and only $1\,^{o}C$ at 6 m depth. The temperature extremes occur about 30 days later at 1 m depth than on the surface; at 6 m depth the

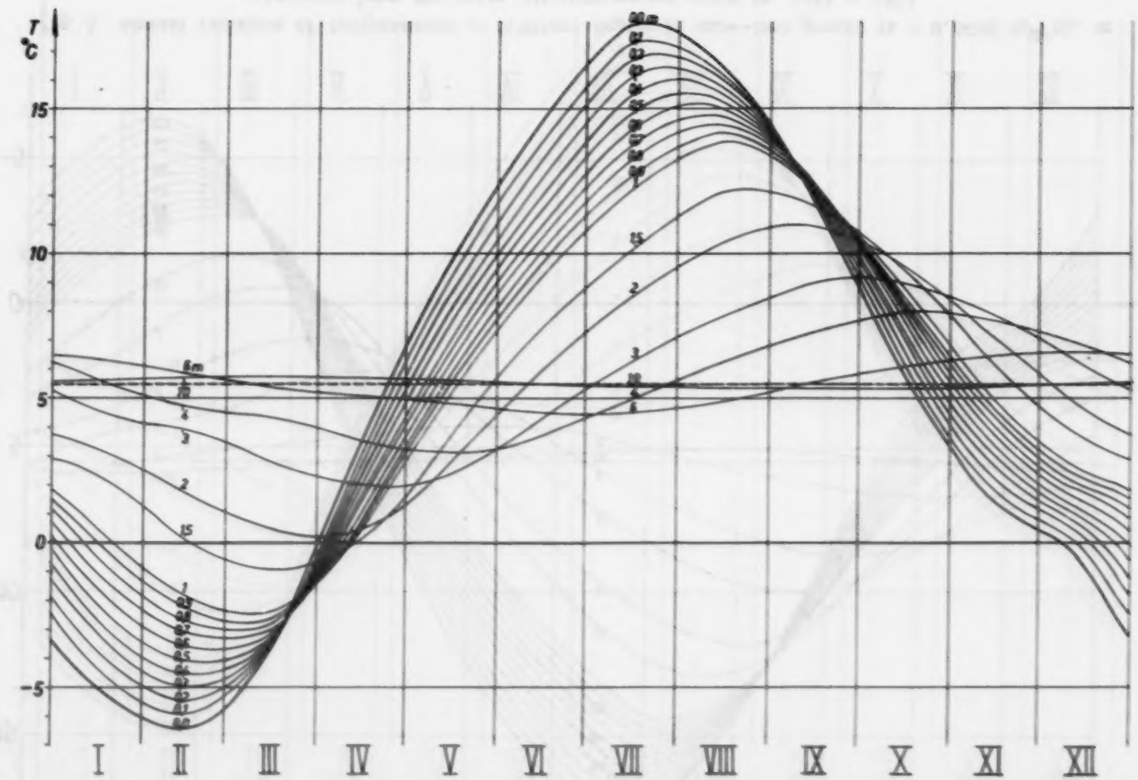

Fig. 4 Annual variation of temperatures at different depths in snow-free ground ($a = 0{,}0022\ m^2/h$), as calculated from the mean air temperature curve in 1945 – 1954

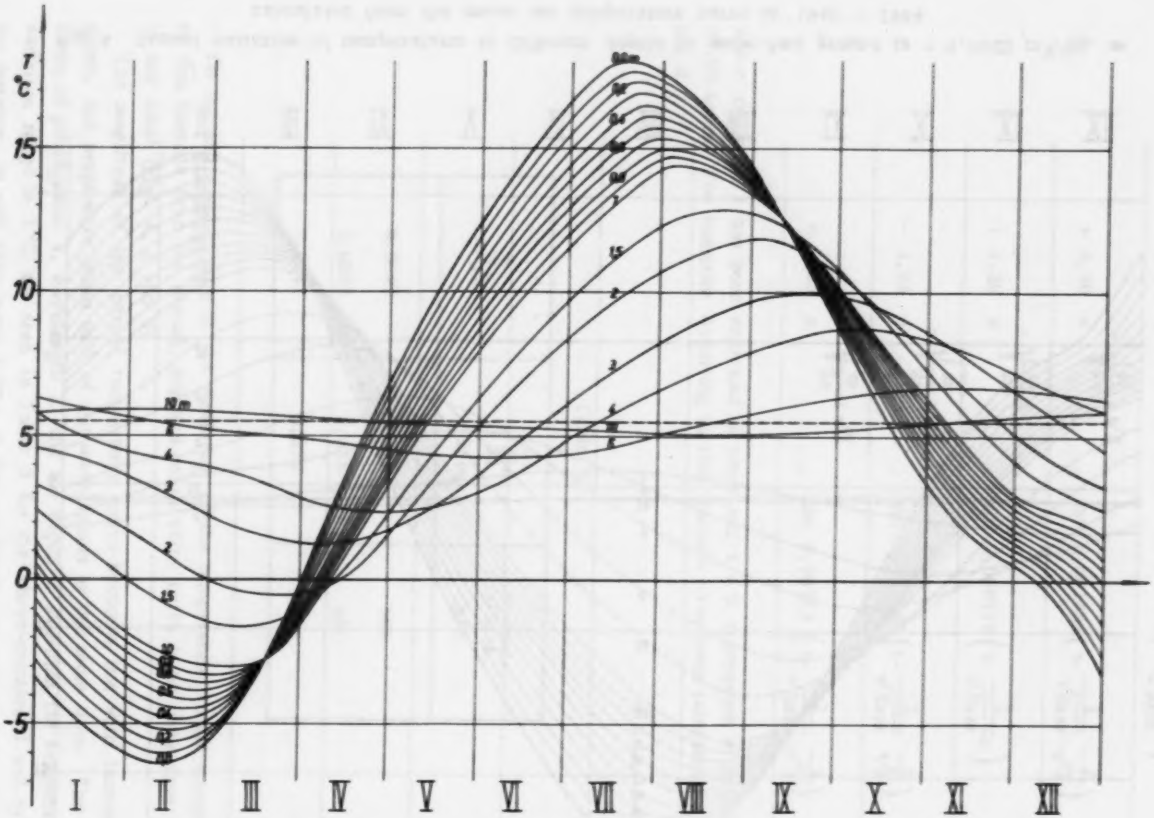

Fig. 5 Annual variation of temperatures at different depths in snow-free ground ($a = 0{,}0032\ m^2/h$), as calculated from the mean air temperature curve in 1945 – 1954

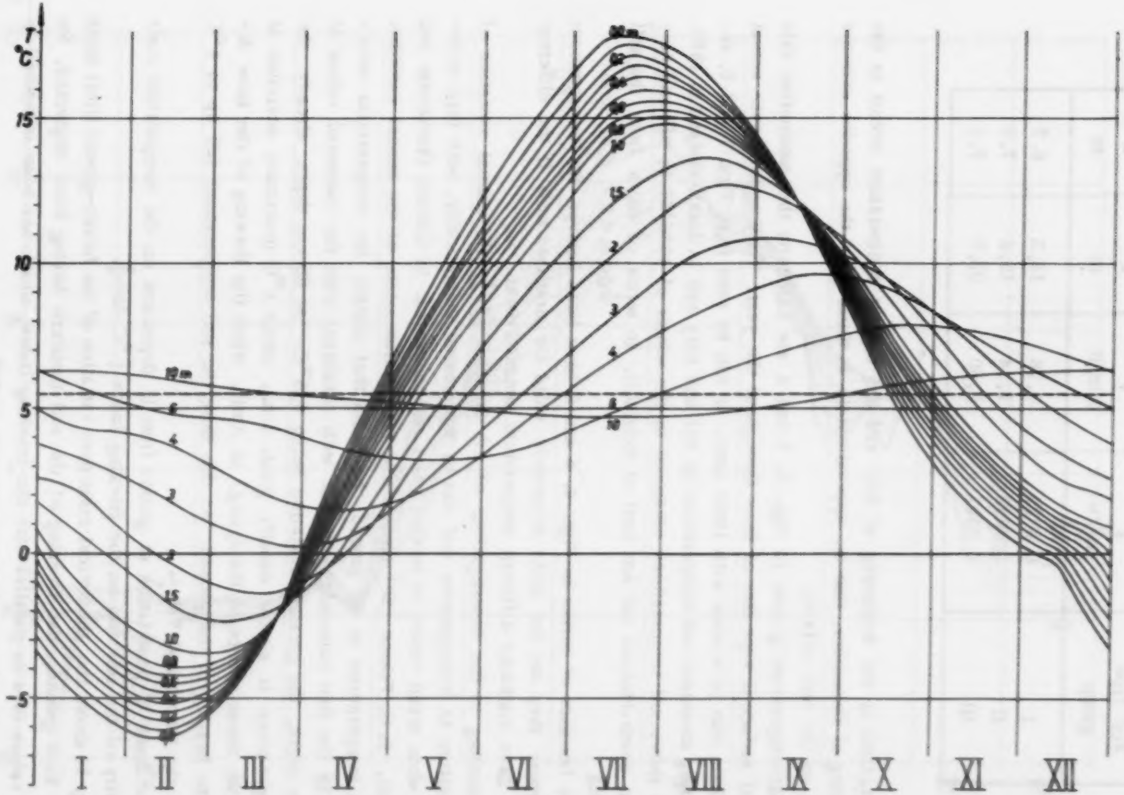

Fig. 6 Annual variation of temperatures in different depths in snow-free ground ($a = 0{,}0047\ m^2/h$), as calculated from the mean air temperature curve in 1945 – 1954

Table 3

Soil type group	a m^2/h	v mm/h	L m	z m
I	0,0022	1,78	15,5	6,2
II	0,0032	2,14	18,8	7,5
III	0,0047	2,60	22,7	9,1

coldest time is the beginning of July and the highest temperature occurs in the beginning of December, that is, approximately at times when the opposite extreme occurs on the soil surface.

The temperature graphs in Figs. 4, 5 and 6 are based on the assumption that the soil surface is kept free of snow throughout the year. They approximate actual conditions also in winters with little snow, as can be seen from Figs. 7 and 8, representing measured soil temperatures in till and clay soil in South-Finland in 1957. Fig. 9 shows soil temperatures measured in the winter of 1956/57 in till and clay soil in South-Finland (10 km west of Helsinki), in terms of depth from the soil surface.

The families of curves in Figs. 4, 5 and 6 have been derived for homogeneous soil types; they are not quite accurately valid for stratified soils if the different layers have markedly different temperature conductivities.

According to the investigations of Ad. Schmidt [19], the annual variation of temperature in homogeneous soil can be represented theoretically, with high accuracy, when mean values of several years are concerned. In Canada (Saskatoon and Ottawa), D. C. Pearce [20] has carried out measurements of the annual variation of the temperatures in the ground; he found that already the temperatures calculated by the first harmonic wave are well consistent with the measured values at 2,4 m depth, the greatest deviation being 0,5 °C. At 60 cm depth, already, the correspondence is almost equally good, i. e., about 1 °C maximum deviation at all other times except in the spring, in April, when the thawing of the snow delays the heating of the ground. At this time the soil temperatures fall up to 4 °C below the theoretical value.

The depth of penetration of ground frost is dependent on the temperature conductivity of the soil and on the freezing index [°C days].

Fig. 10 shows the theoretical maximum variation of the frozen-ground limit depth under bare ground, the freezing of the soil moisture having been neglected, for three values of a in parallel with the freezing index, which has been computed as a mean for the above-mentioned ten-year period.

The variation of the frozen ground limit in snow-covered ground, measured on a small moraine hill and on a field with clayey base soil in the previously men-

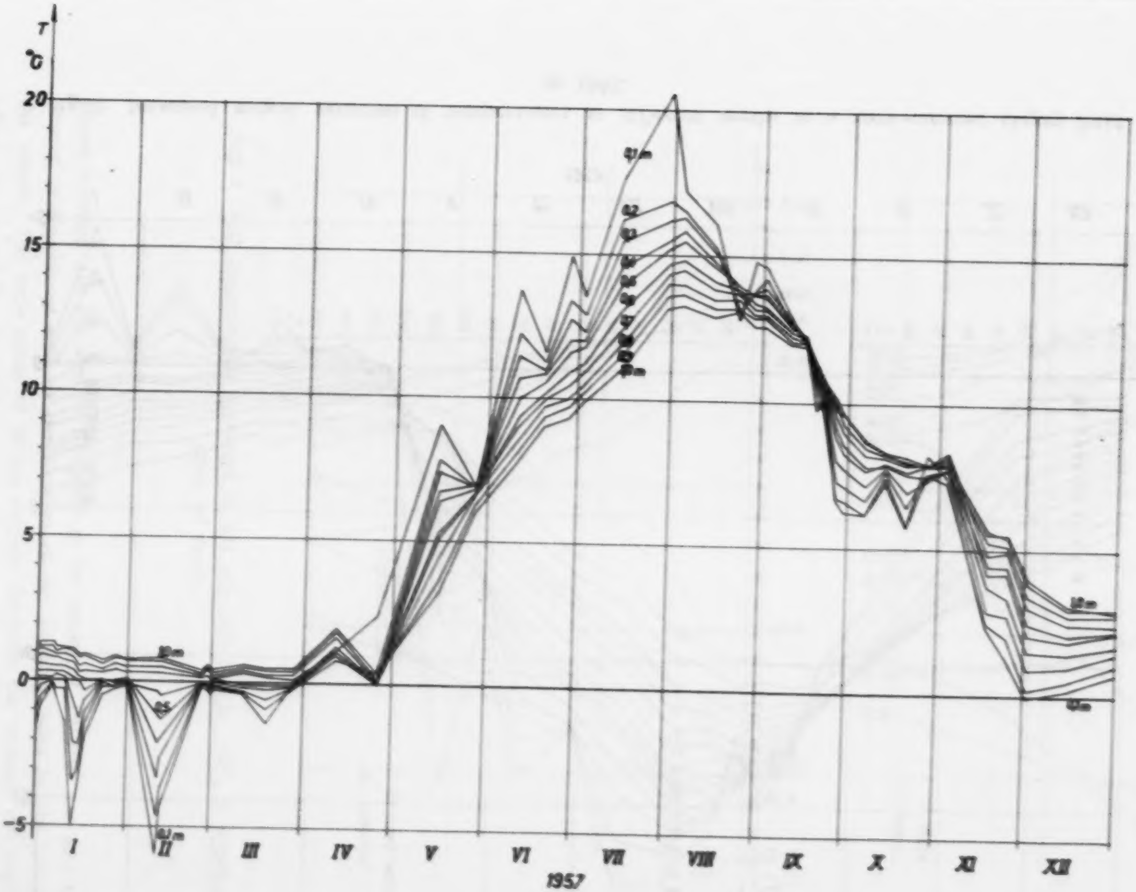

Fig. 7 Measured annual variation of temperatures at different depths in the snow-covered ground of a moraine hill in 1957.

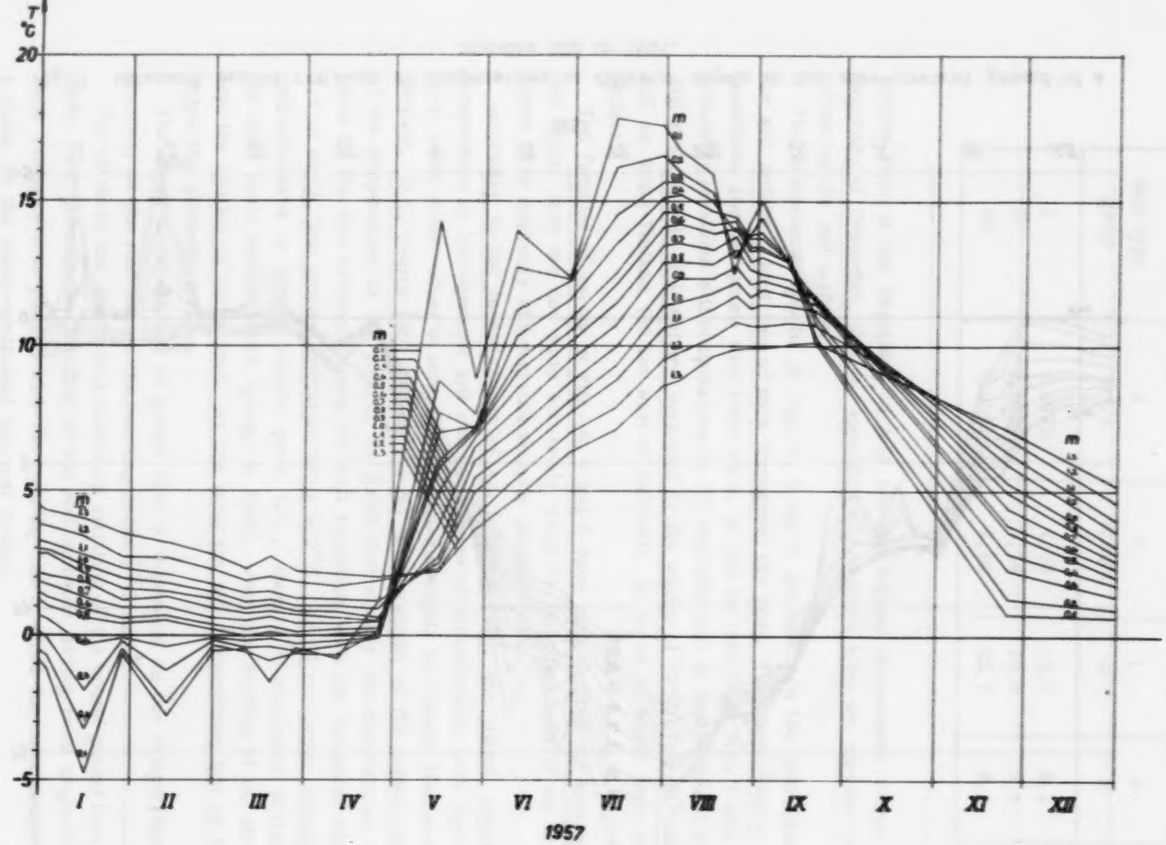

Fig. 8 Measured annual variation of temperatures at different depths in a snow-covered clayey field in 1957.

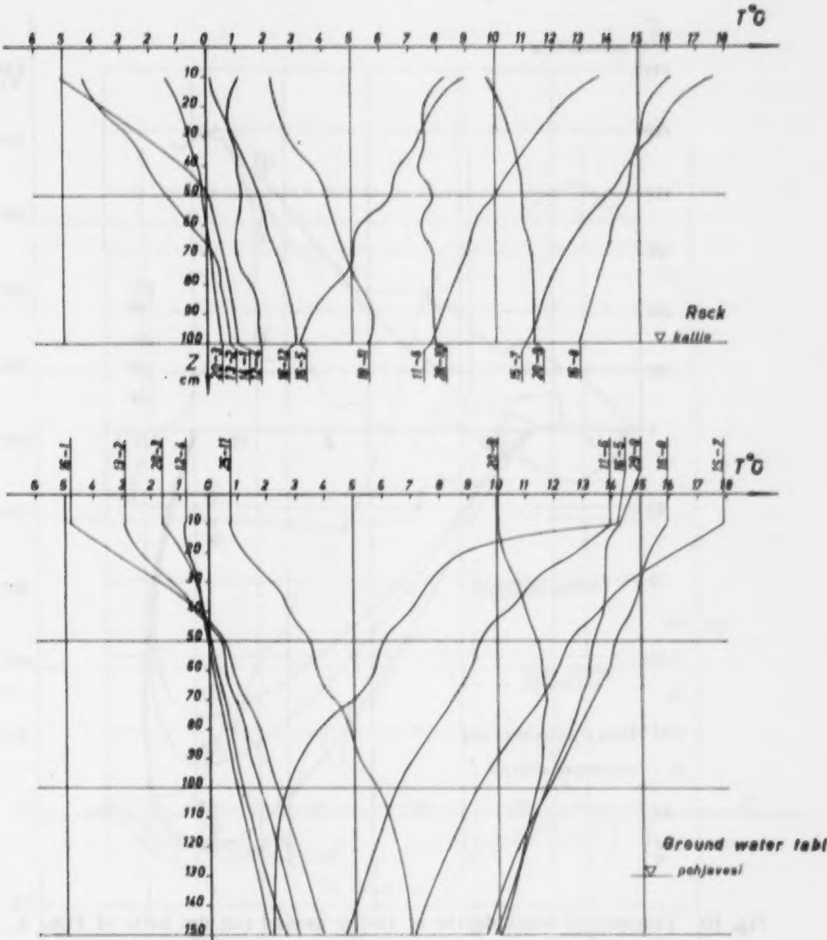

Fig. 9 Soil temperatures measured in till and clay soil in 1957, in dependence of depth.

tioned location (10 km west of Helsinki) in the winter of 1956/57 has also been plotted in Fig. 10, as well as the increase of the freezing index in the course of the winter. The snow cover of 30 cm average thickness has reduced the depth to which the ground froze to about one third of the theoretical value applying to bare ground.

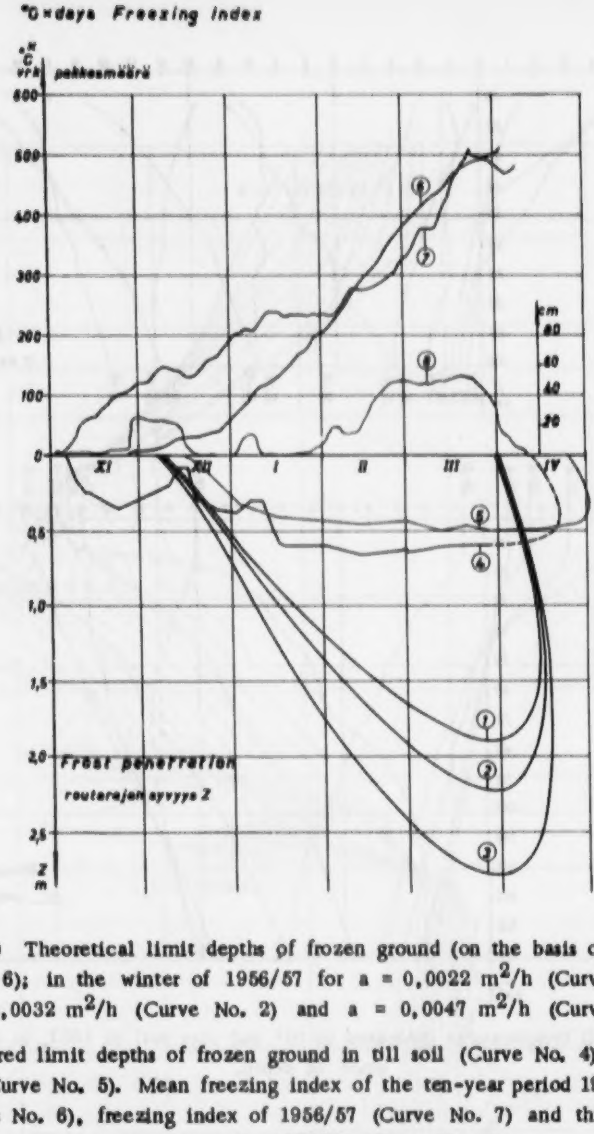

Fig. 10 Theoretical limit depths of frozen ground (on the basis of Figs. 4, 5 and 6); in the winter of 1956/57 for a = 0,0022 m^2/h (Curve No. 1), a = 0,0032 m^2/h (Curve No. 2) and a = 0,0047 m^2/h (Curve No. 3). Measured limit depths of frozen ground in till soil (Curve No. 4) and clay soil (Curve No. 5). Mean freezing index of the ten-year period 1945 ··· 1954 (Curve No. 6), freezing index of 1956/57 (Curve No. 7) and thickness of snow cover in 1956/57 (Curve No. 8).

Measurements of the depth to which the ground freezes have been carried out in Ottawa (Canada) in the winter of 1950/51 in various soil types in snow-covered and bare ground (Fig. 11) [21]. The figure also shows the variation in thickness of the snow cover and the freezing index.

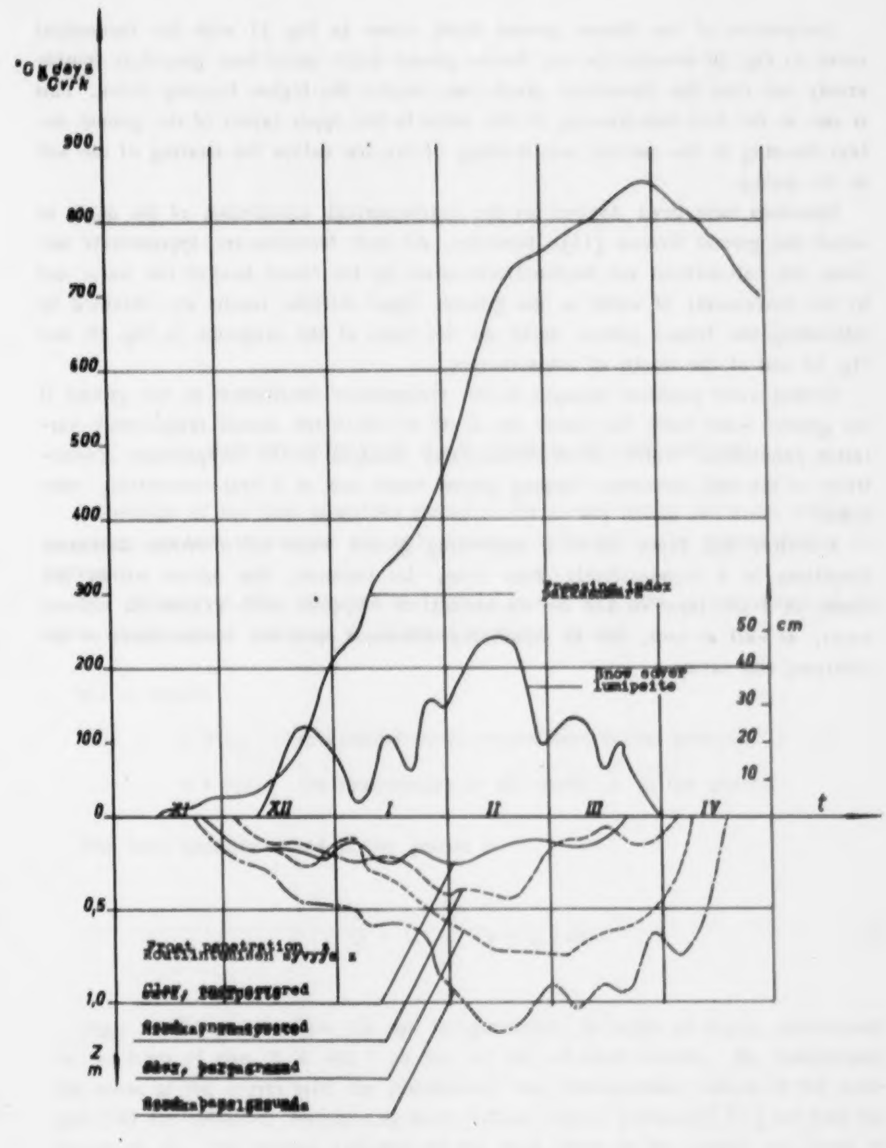

Fig. 11 Variations of limit depth of frozen ground, thickness of snow cover and freezing index in Ottawa (Canada) in the winter of 1950/51.

Comparison of the frozen ground depth curve in Fig. 11 with the theoretical curve in Fig. 10 reveals that the frozen ground depth under bare ground is considerably less than the theoretical prediction, despite the higher freezing index. This is due to the fact that freezing of the water in the upper layers of the ground delays freezing in the autumn and thawing of the ice delays the heating of the soil in the spring.

Equations have been derived for the mathematical calculation of the depth to which the ground freezes [15]. However, all such formulas are approximate because the calculations are rendered inaccurate by the latent heat of the water and by the movements of water in the ground. More reliable results are obtained by estimating the frozen ground depth on the basis of the diagrams in Fig. 10 and Fig. 13 and of the results of measurement.

Ground water produces changes in the temperature distribution in the ground if the ground water table lies above the depth to which the annual temperature variation penetrates. Water causes considerable changes in the temperature conductivity of the soil. Moreover, flowing ground water acts as a heat-transporting medium.

A shallow soil layer above a high-lying ground water table attains stationary conditions in a comparatively short time; for instance, this occurs within one month in a soil layer of 130 cm thickness if $a = 0,0022\ m^2/h$ (Table 3). Ground water, as well as rock, has an equalizing influence upon the temperatures of the overlying soil layers.

4. THE QUANTITY OF HEAT STORED IN THE GROUND

Knowledge of the heat quantities stored in the ground during the warm season is necessary, as they determine the temperature conditions in the ground during the winter.

The quantity of heat stored in the ground at any time can be calculated if the annual temperature variation at different depths is known.

Let us denote

$t = t_0$, the annual mean temperature in the ground,

$t = t(z)$, the temperature at the depth z in the ground.

The heat quantity stored in the ground is

$$Q = C \int_0^{\infty} (t - t_0)\, dz \qquad (15)$$

Figs. 12, 13 and 14 show the soil temperatures, in terms of depth, determined on the basis of Figs. 5, 6 and 7 on the 1st day of each month. By determining the areas of the curves with the planimeter, the corresponding values of the integral (15) are obtained; multiplying them further with C $[\mathrm{Mcal/m^3\ ^\circ C}]$ we find the values of Q. The annual variation of the heat stores in the ground are shown in Fig. 15 for three different values of the temperature conductivity. The heat quantities have been calculated with reference to the mean temperature of the previously mentioned ten-year period, including depths up to 10 m. The maximum variation of the heat stored in the ground, or the total heat exchange of the ground has been shown in Table 4 for bare ground and for the different soil types.

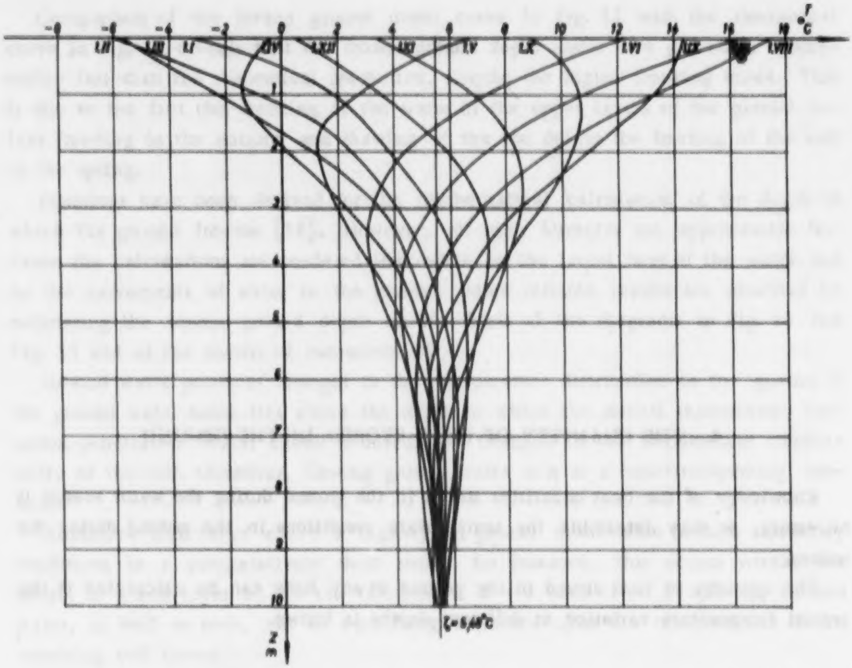

Fig. 12 Temperatures at different depths in bare ground on the 1st of each month according to Fig. 4; $a = 0,0022$ m^2/h.

Table 4

a m^2/h	λ kcal/mh ^oC	C Mcal/m^3 ^oC	$Q_{max} - Q_{min}$ Mcal/m^2
0,0022	1,43	0,650	27,7
0,0032	1,90	0,594	30,0
0,0047	2,50	0,532	34,1

The function representing the annual variation of the soil surface temperature (formula (14)) can also be used to calculate approximately, on the basis of the temperature gradient, the quantity of heat that has been absorbed in the soil

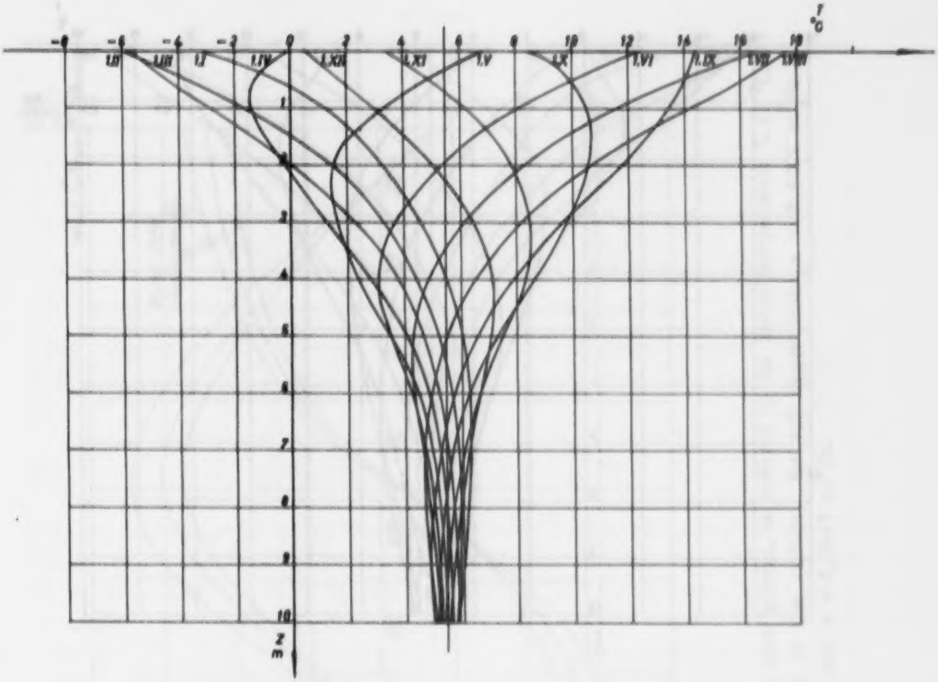

Fig. 13 Temperatures at different depths in bare ground on the 1st of each month according to Fig. 5; $a = 0{,}0032\ m^2/$.

$$Q = -\lambda \int_{t_1}^{t_2} \left(\frac{\partial T}{\partial z}\right)_{z=o} dt \tag{16}$$

The heating of bare ground starts about the middle of March, and the heat stored in the ground is at its maximum in the beginning of September. When $\left(\frac{\partial T}{\partial z}\right)_{z=o}$ is calculated from equation (13), taking into account only its first cosine term, which is graphically represented by Curve No. 3 in Fig. 3, and when this is substituted in equation (16), one finds by integration

$$Q = -\lambda \Delta T \sqrt{\frac{t_o}{2\pi a}} \int_{t_1}^{t_2} \sin\left[\frac{2\pi t}{t_o \pi a} + \frac{\pi}{4}\right] \tag{17}$$

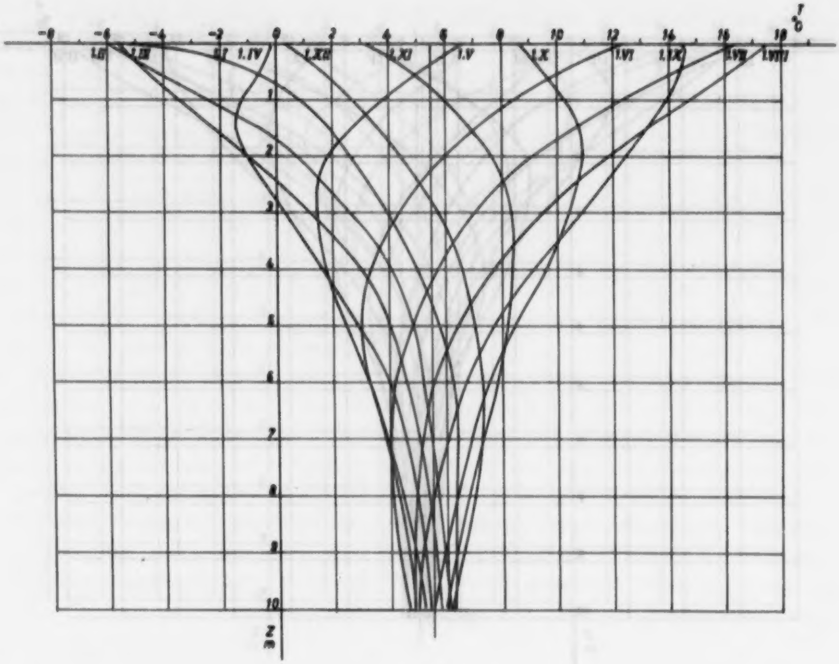

Fig. 14 Temperatures at different depths in bare ground on the 1st of each month according to Fig. 6; $a = 0,0047\ m^2/h$.

Substituting in (17),

$$T = 12,2\ ^{o}C \qquad t_1 = 75 \text{ days}$$

$$t_o = 365 \text{ days} \qquad t_2 = 250 \text{ days}$$

the heat quantity stored in the ground is obtained for the different soil types (Table 5). The heat quantities are 10 ··· 15 % lower than the values in Table 4, which were calculated by a more accurate procedure.

According to different investigators, the total heat exchange of the ground is 15 ··· 30 Mcal/m^2, depending on conditions, calculated as the difference of the maximum and minimum heat quantities contained in the ground, when the heat capacity varies between 0,3 and 0,5 Mcal/m^3 ^{o}C. Most nearly equivalent to conditions in South-Finland are the measurements performed by Luboslavsky [22] in the vicinity of Leningrad. His calculations are based on mean temperature values

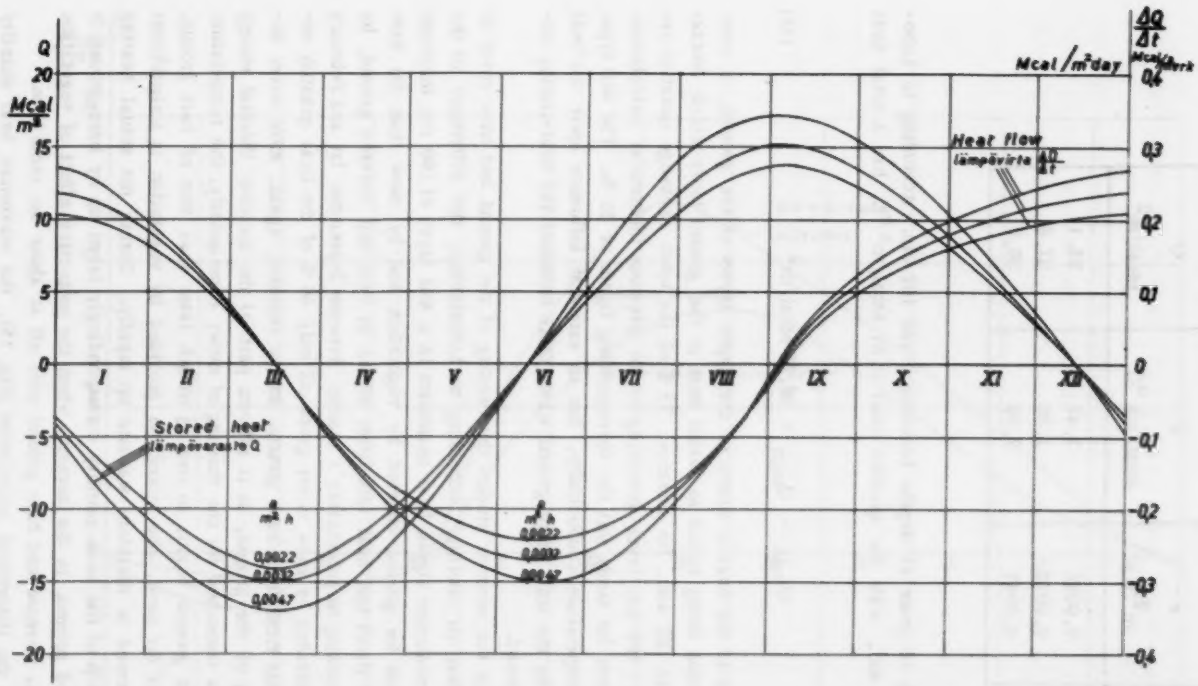

Fig. 15 Annual variation of the heat quantity stored in bare ground, as referred to the mean temperature 5,48 °C of the ten-year period 1945 – 1954; and heat flow to the surface, for soil types with a = 0,0022 m^2/h, a = 0,0032 m^2/h and a = 0,0047 m^2/h.

Table 5

a m^2/h	λ $kcal/mh\ ^oC$	Q $Mcal/m^2$
0,0022	1,43	25,1
0,0032	1,90	27,6
0,0047	2,50	30,0

measured during 15 years at depths between 0 and 160 cm. According to Luboslavsky, sandy soil, with the specific heat 0,57 $Mcal/m^3\ ^oC$, has a total heat exchange of

$$Q_{max} - Q_{min} = 26,51\ Mcal/m^2 \tag{18}$$

The greater part of the heat is stored in the upper layers of the ground, a considerable proportion being bound as latent heat in the ground layer which freezes in the winter. In till soil, for instance, 73 % of the total exchange quantity are stored in the 0 ··· 160 cm layer according to the previous theoretical calculation. Luboslavsky reports for sandy soil the corresponding figure of 61 %. The soil type, especially its temperature conductivity, has an essential influence upon the heat quantity stored by the soil. High ground water table increases the heat-storing capacity of the ground.

Plant cover in the summer hinders the heating of the ground and snow cover in the winter hinders its cooling. According to Luboslavsky, the difference of the maximum and minimum stored heat quantities in a soil layer of 160 cm thickness is about 33 % less for ground covered by vegetation and by snow than for bare ground. Fig. 16 shows the heat quantities stored in bare and covered ground, by the seasons, according to Luboslavsky's results. Between September 1st and February 28th, ground protected by snow cover gives off only 46 % of the heat quantity escaping during this time from bare ground. In the summer, again, snow cover delays the heating of the ground, as it reflects part of the incident thermal energy and part of it is consumed for the thawing of snow; consequently, the temperature in snow-covered ground begins to rise 1½ month later than that of bare ground. Upon thawing of the snow, the protection provided by vegetation is insignificant yet, and the ground is therefore warmed up rapidly. During the actual heating period about 79 % of the heat quantity correspondingly taken up by bare ground is stored in covered ground. In the autumn, when the protective effect of vegetation has become less, covered and bare ground cool off at about the same rate.

According to the theoretical calculation (Fig. 15), the maximum heat quantity is stored in the ground at the beginning of September when the heating period of

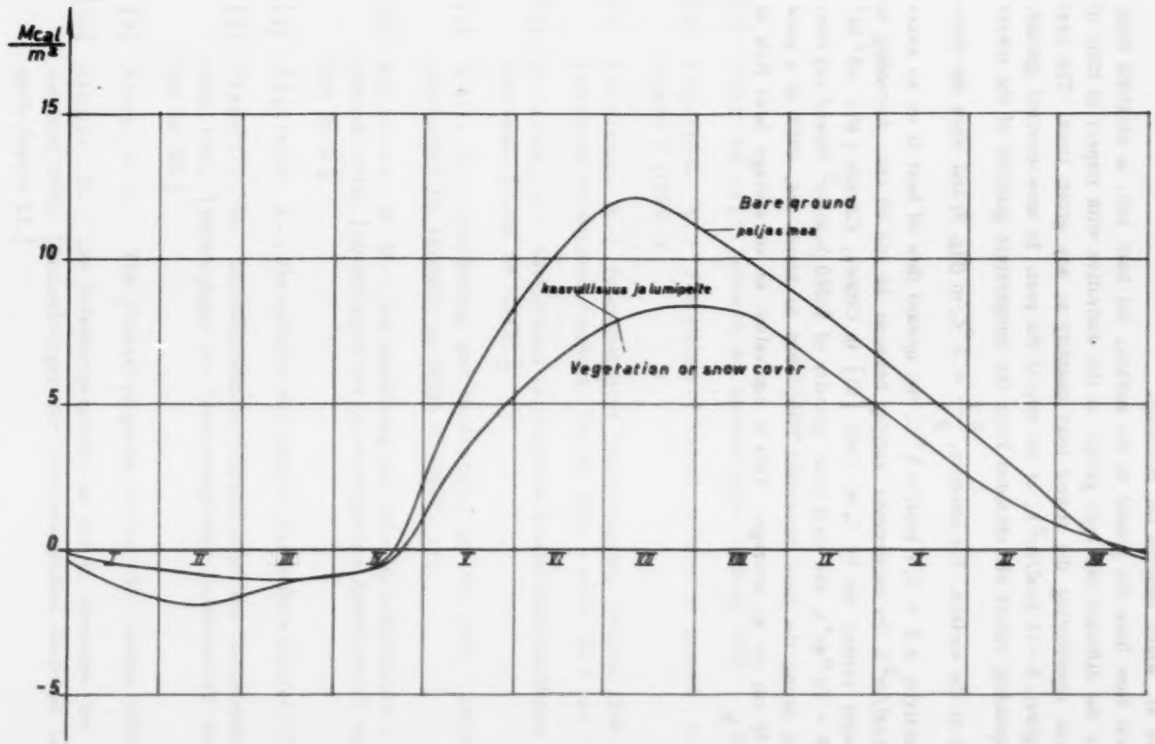

Fig. 16 Annual variation of the heat quantity stored at depths from 0 to 160 cm in bare ground and in ground covered by vegetation in the summer and by snow in the winter, according to Luboslavsky.

houses begins. The heat stores of the ground are at minimum about the 20th of March. During the interval the ground has given off the entire stored heat quantity (Table 4), which benefits the buildings.

The heat flow from the ground to the surface, for bare soil, is obtained from Fig. 15 for the different soil type groups as the derivative with respect to time of the function representing the stored heat quantity at any given time. The heat flow is highest, 8 – 11 kcal/m^2 h, at the turn of the year. In snow-covered ground, the corresponding values are obtained from the temperature gradient of the measured field at the surface. For instance, $\frac{\partial T}{\partial z} = 3\ ^{\circ}C/m$ (Fig. 9) and when the thermal conductivity is $\lambda = 1{,}9$ kcal/m h $^{\circ}$C, the upward flow of heat is on an average 5,7 kcal/m^2 h for snow-cover varying between 40 and 60 cm. According to measurements carried out by L. W. Cold [23] in Ottawa, Canada ($\varphi = 45^{\circ}24'$ N. lat., $\Psi = 75^{\circ}43'$), the total heat quantity of 24500 kcal/m^2 flowed out from the ground during the time November 17th, 1956 to March 3rd, 1957 at a snow depth of 20 cm on an average. This is equivalent to an average heat flow of 9 kcal/m^2 h.

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Page	Line	In place of	Read
15	8	$T(z,t) = T_0 + \sum_1^{\infty} e^{-\sqrt{\frac{n\omega}{2a}}z}\left[A_n \cos n\omega(t - \frac{z}{\sqrt{2a\, n\omega}}) + B_n \sin n\omega(t - \frac{z}{\sqrt{2a\, n\omega}})\right]$	$T(z,t) = T_0 + \sum_1^{\infty} e^{-\sqrt{\frac{n\omega}{2a}}z}\left[A_n \cos n\omega(t - \frac{z}{\sqrt{2a\, n\omega}}) + B_n \sin n\omega(t - \frac{z}{\sqrt{2a\, n\omega}})\right]$
27	1	Fig. 9 Soil temperatures measured in till . . .	Fig. 9 Soil temperatures measured in moraine . . .
28	1	Fig. 10 Theoretical limit depths of frozen ground (on the basis of Figs. 4, 5 and 6); in the winter of 1956/57 for a = 0,0022 m^2/h (Curve No. 1), a = 0,0032 m^2/h (Curve No. 2) and a = 0,0047 m^2/h (Curve No. 3). Measured limit depths of frozen ground in till soil (Curve No. 4) and clay soil (Curve No. 5). Mean freezing index of the ten-year period 1945 – 1954 (Curve No. 6), freezing index of 1956/57 (Curve No. 7) and thickness of snow cover in 1956/57 (Curve No. 8).	Fig. 10 Theoretical limit depths of frozen ground (on the basis of Figs. 4, 5 and 6); for a = 0,0022 m^2/h (Curve No. 1), a = 0,0032 m^2/h (Curve No. 2) and a = 0,0047 m^2/h (Curve No. 3). In the winter of 1956/57 measured limit depths of frozen ground in till soil (Curve No. 4) and clay soil (Curve No. 5). Mean freezing index of the ten-year period 1945 – 1954 (Curve No. 6), freezing index of 1956/57 (Curve No. 7) and thickness of snow cover in 1956/57 (Curve No. 8).

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